MID-LATITUDE AMAZONIAN GLACIATION ON MARS: CONTROLS ON ACCUMULATION AND GLACIAL FLOW PATTERNS. James L. Fastook[1] and James W. Head[2], [1]University of Maine, Orono, ME 04469, fastook@maine.edu, [2]Brown University, Providence, RI 02912.

Introduction: A wide range of evidence shows that the current distribution of ice on the surface is anomalous, and that the Amazonian period was characterized by a variety of non-polar ice-related deposits ranging from *high-latitude* mantles, to *mid-latitude* lobate debris aprons, lineated valley fill, concentric crater fill, and pedestal craters, to *low-latitude* tropical mountain glaciers [1, 2, 3]. General circulation models (GCM) and glacial flow models illustrate the orbital parameter and atmospheric/surface conditions under which periods of glaciation are favored (e.g., [4, 5]), and the resulting patterns of accumulation of snow and the flow of ice [6, 7, 8].

Geological observations and impact crater size-frequency distribution data strongly suggest that during the Late Amazonian, a significant part of the mid-to-high latitudes in both hemispheres was covered by regional snow and ice deposits (preserved today beneath pedestal craters [9. 10, 11]) and that local depressions (primarily impact craters) were the sites of significant ice accumulation, and preservation beneath a residual debris cover (concentric crater fill (CCF) [12]). Pedestal crater (Pd) heights show that a significant amount of snow and ice accumulated in the mid-to-high latitudes during these periods (regionally the mean height is ~50 m, but values up to 160 m are seen in Utopia [13]). Accumulations in CCF are typically many hundreds of meters and can exceed several kilometers [12], filling the crater completely. Could these landforms signify a sufficient thickness of ice to produce active glaciers that flowed across the surface, filling existing lows such as impact craters?

These issues are important with regards to the case of “typical landforms,” for instance, impact craters with concentric-crater fill, CCF. Important questions to answer about what happens when a uniform ice thickness is spread over a depression such as a crater include knowing where flow initiates (is it on the crater rim crests and walls where the topographic slopes are high?) and whether it then flows in from the outside until the crater has filled with ice and the slopes are too low and the flow stops. The answer is certainly yes to both of these, but the timing depends strongly on the temperature-dependent viscosity of the cold ice. Additionally we need to know whether this is still an active process, what the landscape might look like during a period of extensive ice cover, and what happens during the period of ice retreat and loss (the waning stages of the glaciation, during which as the ice surface lowers and the crater rim crests are exposed, rocky debris can be added to the inner slopes of the crater providing the source of the debris cover on the surfaces of the CCF).

All of this relates to understanding the general characteristics and dynamics of extensive Amazonian glacial periods in the mid-latitude of Mars. It is critical to differentiate between the possibilities that there were accumulation-driven regional ice sheets whose flow was accentuated at steep topographic slopes or whether there were only thinner regional mantles of snow and ice that only flowed within craters or at other areas with steep topographic slopes.

Discussion: Concentric Crater Fill (CFF) is a morphologic formation observed within a relatively narrow middle to high latitude band in both the northern and southern hemispheres of Mars. In its most ”classic” form, CCF is a crater-interior unit with concentric lineations and topographic ridge and troughs (as many as eight), typically a few hundreds of meters wide, an example of which is shown in Fig. 1 (left from [14]). Classic CCF is characterized by a surface texture dubbed “brain terrain,” a sample of which is also shown in Fig. 1 (right from [13]).

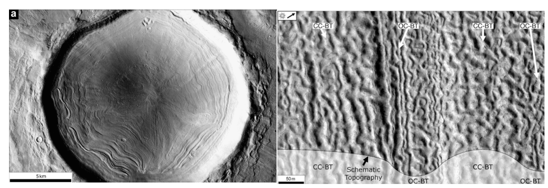

Figure 1: A crater with CCF (left from [14]) and characteristic “brain terrain” (right from[13]).

Brain terrain [13] is thought to be the result of atmospheric deposition of ice at high obliquity, followed by cold desert modification that includes glacial flow deformation, thermal contraction cracking, differential sublimation and no liquid water. These fracture networks orient to reflect and preserve evidence of the flowing stress field. Enhanced sublimation widens the cracks and sand wedges form, which later protect the ice resulting in depressions on the polygon surface as the ice-rich layer sublimates. Brain terrain is also observed on Lineated Valley Fill (LVF) and Lobate Debris Aprons (LDA), features of similar age and provenance as the CCF. CCF does not merely line or coat the interior [14], but can manifest as high and flat or even concave-up deposits that fill the crater to depths of 600 to 1700 m, as much as 75-80% of the original crater depth [14, 15]. An example profile across a CCF-crater is shown in Fig. 2. In addition, transgressions of the crater rim (Fig. 3) indicate that the surface had to have been 300 m higher at some point in the past.

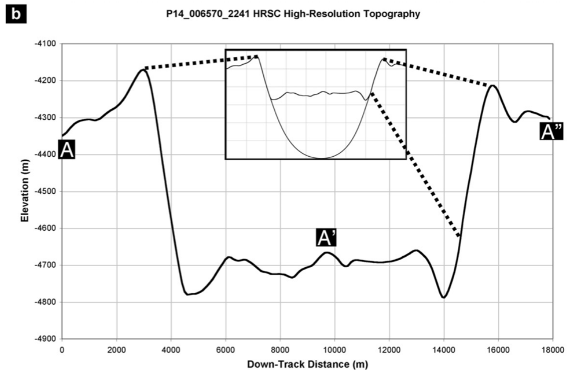

Figure 2: HRSC high-resolution DTM profile across CCF-containing crater with inset showing expected depth from [15].

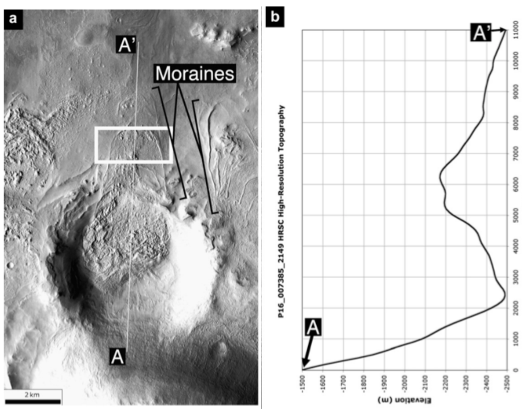

Figure 3: a) CCF-crater showing flow across the crater rim. b) HRSC high-resolution DTM profile across crater and nearby plains from [14].

Various hypotheses have been proposed to explain CCF. These include aeolian processes [16], debris from bedrock recession of scarps [17], ice-assisted talus flow [18], rock glaciers [19, 20, 21], internally deforming ice [22], and debris-covered glaciers [13, 23]. Levy et al. [14] discusses in detail the different arguments and arrives at the conclusion that CCF is composed largely of debris-covered ice that at some point flowed and then sublimated. They also conclude that the CCF is likely Amazonian in age, reflecting processes that have taken place within the last 320 Myr.

Levy et al. [14] suggest that accumulation in the alcoves could possibly be feeding the CCF. They also suggest, however, that the surface of the CCF, which has lowered by as much as 300 m, could have overfilled some of the craters, so perhaps the CCF material could have been part of a larger inter-crater ice complex that covered the entire landscape. As such, the CCF may be the remnant of that larger ice complex.

A point of contention in the debris-covered-ice argument is at what thickness does ice flow in the Amazonian climate of Mars. The hardness of ice at typical Martian temperatures is in fact the centerpiece of the argument for aeolian emplacement [16]. Others [8] have suggested that some limited flow is possible even at Martian temperatures. Both of these treatments dealt only qualitatively with the resulting flow deformations. Our ice sheet model includes a temperature-dependent ice rheology, and so a true quantitative assessment within the context of an ice sheet model is possible. We explore the nature of regional ice accumulation and glaciation during this time period and address the following questions:

1) What thickness is required to initiate ice flow on a flat, inter-crater area under Late Amazonian conditions? Velocity is obtained by integrating strain rates through the vertical, with strain rates related to stresses through the Flow Law with a temperature-dependent rate factor [24]. Flow velocity depends on surface slope cubed and ice thickness to the fourth power. Fig. 4 shows $\log_{10}$ of the flow velocity as a function of ice thickness and surface slope for temperatures of 215, 225, and 235 K. At 215 K the low surface slopes (~1^o) typical of the flat inter-crater terrain require considerable thickness (800-1000 m) to initiate flow. Only for considerably larger surface slopes (>5^o) is there any significant flow for 200 m. Even at slopes as high as 20^o, 50 m only yields 0.3 mm/yr. A temperature of 225 K increases all velocities by a factor of 4, reducing the threshold for 200 m to a slope of 3.5^o. A further increase to 235 K reduces the necessary slope to 2.2^o, but 100 m thick ice still requires a slope greater than 5^o to generate a flow velocity of 1 mm/yr.

2) Could the current Pd mean thickness value (~50 m) be the remnant of equilibrium flow (that is, did thicker ice flow until it reached an equilibrium thickness similar to the current observed Pd thickness)? Was the Pd layer the last phase of a thick persistent ice sheet that reached a configuration that supported flow? Or was it a transient, relatively thin ice-rich layer that deformed as it covered and flowed into the crater depressions?

Evidence exists for the latter case in the form of Pd, perched craters, and excess-ejecta craters described in detail in [9, 10, 11]. These three types of craters relate to the impacts into an ice-rich layer that is at most a few hundred meters thick. Each type reflects differing degrees of penetration, followed by complete sublimation of any un-armored regions of the ice complex.

Repeated deposition and removal of this thin layer is suggested in that ~80 m height difference is observed between two Pds 20 km apart. In addition, 30 have superimposed Pds. For this to occur, the first Pd would have to form, the entire ice-rich layer outside the armored zone would have to be removed, a second ice-rich layer would have to reform, and the superimposed Pd could then be emplaced. Clearly this requires that there be multiple episodes of ice-rich layer cover. GCM results [4, 5] predict ice accumulations as high as 10 mm/yr during periods of high obliquity exactly in the mid-high latitudes where these craters occur.

One might ask why the transient layer never gets any thicker than the 50-160 m suggested by the Pds, since GCM results suggest this layer could form in as little as 20 Kyr, a time much shorter than the duration of the obliquity excursions. Estimates of the volume of the Pd-defined layer [9,10, 11] are close to the known volumes of the polar caps that are the source of moisture for the high-obliquity mid-high latitude precipitation. The transient layer is "supply-limited" in that when the cap is exhausted, the source is removed, and deposition ceases even if the obliquity is still high.

3) What slopes are required to initiate ice flow under Late Amazonian conditions and where is this most likely to occur geologically? Evidence exists for a wide-spread transient ice-rich layer 50 to a few hundreds of m thick in the mid-high latitudes where GCM results deposit ice at high obliquity. How does this thin layer become the several-hundred-meters- to kilometers-thick deposit observed as CCF? As we described, significant flow of thin ice at Amazonian temperatures can only occur for relatively steep slopes. Garvin et al. [15] show that crater-wall slopes correlate strongly with crater size, with slopes from 10-30^{o}. Fig. 4 shows that these slopes would easily provide significant flow, even for layers less than 200 m.

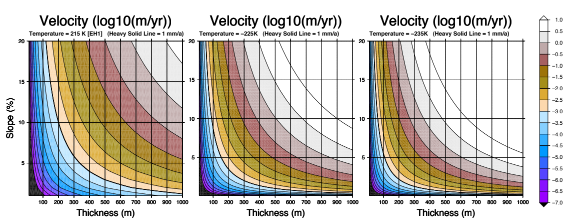

Figure 4: Flow velocity (log_{10}(m/yr) as a function of ice thickness (m) and surface slope (%) for a temperatures of 215, 225, and 235 K with the heavy solid line in each figure indicating a velocity of 1 mm/yr.

A transient layer that uniformly blankets the terrain and flows down the steep walls into the crater interior thickens the deposit there, which is then less likely to completely sublimate during the next episode of low obliquity. Also the re-exposed crater walls provide a source of debris that can armor the crater-interior ice surface, adding to its likelihood of surviving until the next cycle of high obliquity. Layer formation and removal must happen many times, with the layer repeatedly forming, flowing, and sublimating away. In the obliquity-driven movement of water to and from the mid-high latitudes, a small amount is deposited in the crater depressions in each cycle, accumulated there by flow down the steep slopes of the crater walls.

4) What was the nature of ice cover and glaciation during periods of maximum ice accumulation in the late Amazonian? What can we say about the possible inter-crater ice complex? Was it a thick persistent ice sheet? Or was it a transient relatively thin ice-rich layer that deformed as it covered and flowed into the crater depressions? Evidence exists for the latter case in the form of pedestal craters (Pd), Perched craters (Pr), and excess-ejecta craters (EE) described in detail by Kadish [9, 10, 11]. In these papers Kadish relates the three different types of craters to the impact of meteorites into an ice layer that is at most a few hundred meters thick. Each of the different types reflects differing degrees of penetration into the ice--rich layer, followed by ultimate complete sublimation of any un-armored regions of the ice complex. Detailed dating of the Pds allows Kadish to estimate their formation time as well as how frequently and for how long the ice complex had to be in place for the observed distribution of Pds to be produced.

The best fit formation time for the observed distribution of Pds is ~90 Myr, but this may be a cumulative time, with the actual formation perhaps spread over a much longer time due to coming and going of the ice-rich layer. This coming and going of the ice layer is suggested by the fact that as much as 80 m height difference can be observed between two Pds only 20 km apart. The shortest Pds have a uniform higher density everywhere, whereas the taller Pds occur only at higher latitudes. In addition, over 30 Pds show other Pds superimposed. For this to occur, the first Pd would have to form, the entire ice-rich layer outside the armored zone would have to be removed, a second ice-rich layer would have to reform, and the superimposed Pd could then be emplaced. Clearly this requires that there be multiple waxing and waning episodes of ice-rich layer cover. This is not unreasonable, given that GCM results [4, 25, 26, 27, 28] predict ice-equivalent accumulations as high as 10 mm/yr during periods of high obliquity exactly at the middle to high latitudes where these craters occur (200 m thick layer emplaced in 20 Kyr). Obliquity during the last 5 Myr varies between 15 and 35 degrees, while during the previous 15 Myr it varied from 25 to 45 degrees [29]. It is worth noting that the Laskar solution is only robust for the last 20 Myr, prior to which the solution is chaotic, however similar patterns are observed in many of the possible solutions.

HIRISE/CTX crater counting on the Pd surface [14] yields individual ages from 1 Myr - 3.6 Gyr, with median of 140 Myr, with 70% < 250 Myr old. Of note is the fact that during the period 25 Myr - 175 Myr (a 150 Myr time period) there is at least one Pd emplaced every 15 Myr. The pattern that emerges is that the Pds were likely to have been emplaced into a fluctuating ice layer with a period that is no longer than approximately 15 Myr. Assuming this periodicity, the 90 Myr necessary to form the Pds, and assuming mass balance variation is roughly sinusoidal (half accumulation, half ablation) suggests 180 Myr, (12 cycles).

For this we turn to a 1D flowband model based on the University of Maine Ice Sheet model (UMISM) [6, 30, 31]. UMISM is a shallow-ice model that we have coupled with an advection-based model of debris transport. As the ice surface drops below the rim of the crater, debris is deposited on the ice and transported forward with the flowing ice.

We first perform an experiment to see how long it takes for the crater to fill to a level that matches an observed CCF crater (P14_006570_2241 [14]). We begin with a uniform thickness (200 m) and hold fixed the boundary thickness. A comparison between the model results (black) and the observed CCF crater profile (red) is shown in Fig. 5a. It takes 450 Myr for the model crater to fill to the level in the CCF crater.

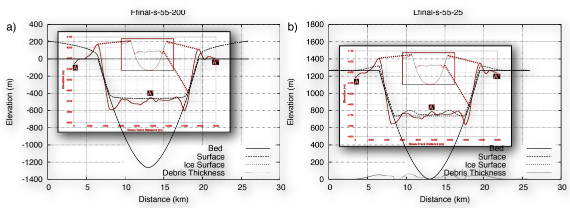

Figure 5: a) Comparison between simulated crater filling from persistent 200 m layer and CCF-containing crater (red, P14 006570_2241 [14]) after 450 Myr, b) simulated crater driven by 301003BIN_A_P001_N obliquity solution [29] and the same CCf-containing crater.

Next we subject this ice sheet/debris model to a climate driven by an obliquity scenario [29] with repeated cycles of ice-layer formation during the time when the Pds formed shown in Fig. 6a. Obliquity calculations are only robust for the last 20 Ma, beyond that the solutions are chaotic (ie. extremely sensitive to initial conditions, but not random). The chosen scenerio is one in which the mean obliquity is relatively high from 40 Ma until 5 Ma, at which point it drops to its current value. Fig. 6b shows a blowup of the transition period from 7.5 to 2.5 Ma. We chose an obliquity threshold of 35^o, above which we have a positive mass balance (1 mm/yr), and below which we ablate the ice. Prior to 5 Ma, the mean obliquity is above this threshold.

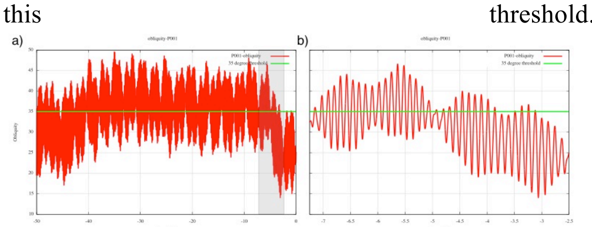

Figure 6: Obliquity for [29] scenario 301003_BIN_A_P001_N a) for the last 50 Ma and b) for the period 7.5 Ma to 2.5 Ma, indicated with grey overlay in a). During this period the mean obliquity shifts from high to low. The horizontal green line shows the 35^o obliquity threshold above which an ice layer is deposited

We limit the deposited layer to the specified thickness by turning off the precipitation when the layer volume has reached a "supply-limited" value. Even at cold temperatures ice is transported into the crater, resulting in thicker ice there and thinner ice on the slopes and inter-crater terrain. With negative SMB, not all of the ice in the crater may be removed and the crater can fill with ice and transported debris. In addition we can reduce the negative SMB as the debris layer armors the ice beneath it. The start-and-stop nature of the forward motion of the ice dictates that the transported debris layer will not be uniform in thickness and it can form concentric ridges similar to those observed in the CCF. These can be seen in Fig. 5b. Note the ripples match the scale and amplitude of the observed surface.

Conclusions: Pds provide us with a means of estimating the thickness of the ice layer that must have periodically mantled the mid-high latutudes of Mars during the Amazonian. Focusing on the CCF as an example of a landform that is glacial in origin, if not also in content, we demonstrate that flow from an inter-crater terrain layer compatible with the Pds measurements cannot fill the craters in the allotted time. We then show how a cyclical pattern of recurring layers, which is in agreement with Pd observations, can both fill the craters with a significant volume of ice as well as transport debris from the crater walls out into the central regions of the craters. The cyclical pattern of waxing and waning mantling layers results in a rippled pattern of surface debris extending out into the crater interiors that would manifest as an observable concentric pattern compatible with the appearance of the CCF. Finally we have driven the simulation with a representative obliquity solution where the layers are assumed to form when obliquity is above a 35^o threshold, helping to determine which of the many chaotic solution might be most likely to have occurred.

References: [1] Head and Marchant 2008 *LPS* 39 #1295. [2] Head et al. 2003 *Nature* 426, 797-802. [3] Head et al. 2006 *Earth and Planetary Science Letters* 241, 663-671. [4] Forget et al. 2006 *Science* 311(5759), 368-371. [5] Madeleine et al. 2009 *Icarus* 203, 390-405. [6] Fastook et al. 2008 *Icarus* 198, 305-317. [7] Fastook et al. 2011 *Icarus* 216, 23-39. [8] Milliken et al. 2003 *J. Geophys. Res.* 108 (E6), doi:10.1029/2002JE002005. [9] Kadish and Head 2009 *J. Geophys. Res.* 114, E10001. [10] Kadish et al. 2010 *Icarus* 210, 92-101. [11] Kadish and Head 2011 *Icarus* 215, 34-46. [12] Kadish and Head 2011 *Icarus* 213, 443-450. [13] Levy et al. 2009 *Icarus* 202, 462-476. [14] Levy et al. 2010 *Icarus* 209, 390-404. [15] Garvin et al. 2002 *LPS* 33 #1255. [16] Zimbelman et al. 1989 *LPS* 19 397–407. [17] Sharp 1973 *J. Geophys. Res.* 78, 4073– 4083. [18] Squyres 1978 *Icarus* 34, 600–613. [19] Squyres 1979 *J. Geophys. Res.* 84, 8087–8096. [20] Squyres and Carr 1986 *Science* 231, 249–252. [21] Haberle et al. 2006 *Science* 311, 368-371. [22] Lucchitta 1984 *J. Geophys. Res.* 89, B409–B418. [23] Garvin et al. 2006 *Meteorit. Planet. Sci.*, 1659–1674. [24] Paterson 1994 *The Physics of Glaciers*. Pergamon, Oxford, 3rd edition. [25] Mischna et al. 2003 *J. Geophys. Res.*, 108, 5062, doi: 10.1029/2003JE002051. [26] Levrard et al. 2004 *Nature*, 431, 1072-1075. [27] Mischna and Richardson 2005 *Geophys. Res. Lett.*, 32, L03201, doi:10.1029/2004GL021865. [28] Levrard et al. 2007 *J. Geophys. Res.*, 112, E06012, doi:10.1029/2006JE002772. [29] Laskar et al. 2004 *Icarus* 170, 343-364. [30] Fastook 1993 *Computational Science and Engineering*, 1(1), 55-67. [31] Fastook et al. 2004 *LPS* 35 #1352.